2.4 SUBSURFACE FLOW

Subsurface flow in the SFWMM can be divided into four processes: infiltration and percolation, canal seepage, levee seepage and groundwater flow. Infiltration refers to the vertical movement of water across the land surface and percolation is the recharge to the water table. Canal-groundwater seepage describes the movement of canal water into the adjacent soil (and vice-versa) by virtue of the differences between the hydraulic head in the canal and that of the water table. Levee seepage is a process wherein surface water moves across a levee embankment and ends up on a levee borrow canal (e.g., from WCA-3B to L-30 borrow canal). Regional groundwater flow (or simply groundwater flow) corresponds to the horizontal movement of groundwater after all of the above processes have occurred. The following four subsections describe these processes in greater detail.

Infiltration and Percolation

Infiltration is the process by which water on the soil surface enters the soil. Water may come from rainfall and/or irrigation and increases moisture in the unsaturated zone or directly goes to the saturated zone via percolation. Percolation is the recharge to the saturated zone or the amount of water crossing the water table. In South Florida, where unconfined aquifer conditions exist, the location of the water table determines the upper limit of the saturated zone. Ponding exists when the water table elevation exceeds the land surface elevation and the unsaturated zone no longer exists. Infiltration and percolation are assumed to be vertical processes.

The volume of infiltration is taken as the minimum of the following three quantities:
1. available water (above land surface) to infiltrate;
2. infiltration rate multiplied by grid cell area and time step; and
3. available void space between the water table and land surface.

Infiltration rates vary from grid cell to grid cell and were determined from the soil classification scheme used for the entire model domain. They range from a value of 9 to 100 ft/day.

Percolation is the amount of water that enters the saturated zone when field capacity (maximum moisture content that can be stored in the unsaturated zone) is exceeded.

Canal-Groundwater Seepage

The interaction of canals with the water table can be modeled by quantifying the exchange of surface water (in the canal) and ground water (in the aquifer). Although generally referred to as canal seepage, leakance or leakage, water can actually leave and enter a canal depending on the relative stages of the local groundwater and the canal itself, hence the term canal-groundwater seepage or canal-aquifer interaction. Seepage is added (or subtracted) from the recharge term which goes into the solution of the groundwater flow equations [Eq. (2.4.3)]. The volume of

seepage into or out of the canal to or from the aquifer is calculated at each node where the canal passes through for every time step. Canal-groundwater seepage is given by

$$CGSEEP_{node,\ t+1} = \left(H_{node,\ t} - SWL_{node,\ t}\right) \bullet CHHC_{node} \bullet DT \bullet 1.4 \bullet RCAR_{node} \qquad (2.4.1)$$

where:

$CGSEEP_{node,\ t+1}$ = seepage volume, [L^3];
$H_{node,\ t}$ = water level at the node (grid cell) through which the canal passes, [L];
$SWL_{node,\ t}$ = canal surface water level at the same nodal location as H_{node}, [L];
$CHHC_{node}$ = canal-aquifer conductivity or connectivity coefficient, [L/T per L of head difference];
DT = length of one time step, [T]; and
$RCAR_{node}$ = length of canal within the node multiplied by the width of the canal.

Since RCAR represents the area of the canal bottom, it is necessary to multiply it by a factor of 1.4 in order to approximate the entire bed or wetted area of the canal at the particular node in question, i.e., channel bottom plus side slopes. Seepage is assumed to occur uniformly within the wetted area of the canal. By SFWMM convention, seepage volume is positive if there is inflow to the canal and negative, otherwise. Variable CHHC ranges from 0.01 to 9.00.

Levee Seepage

Levee seepage refers to the movement of groundwater beneath and through a levee, and into the corresponding levee borrow canal or vice versa. Investigations conducted by the Corps and USGS indicated that significant amounts of seepage occur from the Water Conservation Areas across the major levees to the east.

Figure 2.4.1 shows the SFWMM representation of the total groundwater flow beneath a levee. It is the sum of the regional groundwater flow or underseepage (Q_{US}) and levee seepage (Q_{LS}). Prior to version 2.1, groundwater flow was completely characterized in the SFWMM by numerically solving the governing partial differential equation for transient flow in a two-dimensional, anisotropic, heterogeneous, unconfined aquifer. However, the level of discretization (2 miles x 2 miles) available in the model was considered too coarse for modeling local groundwater phenomenon such as levee seepage. The model's solution to the general groundwater flow equations represents regional groundwater flow while an empirical levee seepage equation is used to solve for levee seepage. The levee seepage algorithm in the SFWMM affords great flexibility because it was incorporated without changing the spatial resolution of the model (Brion and Guardo, 1991).

The basis for the empirical equations representing levee seepage in the SFWMM is an independent set of computer simulation runs using a two-dimensional (vertical plane) model, SEEP2D (a.k.a. SEEPN). Developed at the U.S. Army Corps of Engineers Waterways Experiment Station, SEEP2D simulates steady-state subsurface flow through a multi-layered

aquifer system (confined or unconfined) by solving the Laplace equation using Darcy's Law (Tracy, 1983; Biedenharn and Tracy, 1987).

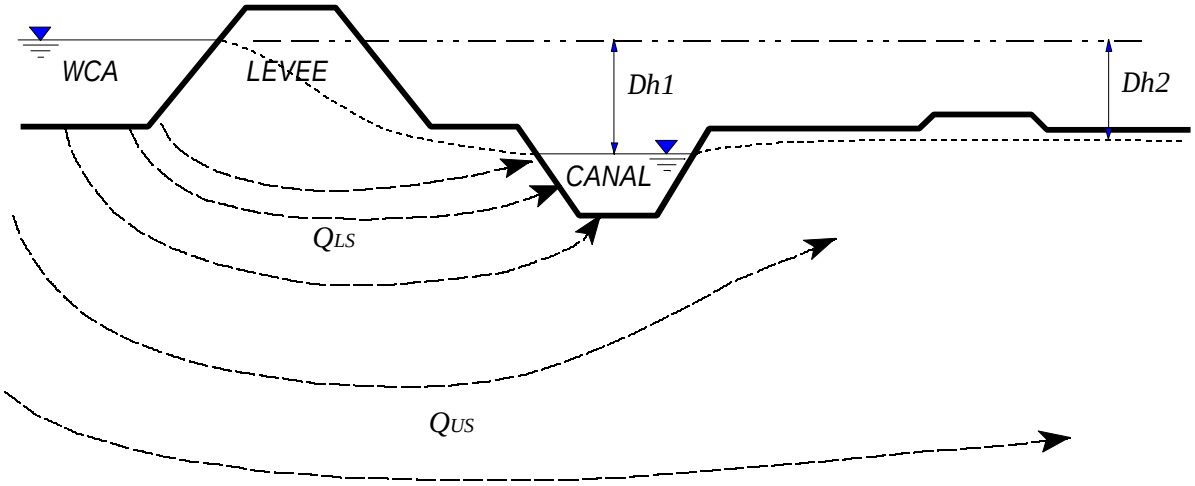

Figure 2.4.1 Canal-Levee Configuration Representing a Typical Transect Used in Developing Empirical Levee Seepage Equations in the South Florida Water Management Model

A concise description of the steps taken in establishing the preliminary empirical levee seepage equations is outlined below:

1. Create a 2-D strip model for each selected (based on similar hydrogeologic characteristics) levee cross-section. Based on Corps general and detailed design memoranda, levee configurations and hydrogeologic properties were compiled and reformatted in accordance to requirements of the SEEPN model. The locations of the sixteen transects (along L40, L36, L35B, L35, L37, L67, L33, L29 and L31N levees) used in this analysis are shown in Fig. 2.4.2.
2. Run SEEPN for different combinations of hydraulic heads and canal stages. Stages in the water conservation areas, borrow canals and areas just east of these canals, that were deemed representative of steady-state conditions (wet, dry, and average) for all transects, were selected as input to the SEEPN model. Model output was summarized to determine the capture rate (amount of total seepage beneath a levee that ends up in the borrow canal) for each model run.
3. Propose empirical equations and derive regression coefficients for the equation relating volume of water captured by borrow canal to total head gradient immediately across the levee (local head gradient) and from cell-to-cell (regional head gradient). Consistent with Darcy's Law, the independent variables in the functional form of the regression equation were chosen as head gradients, instead of absolute stages, since hydraulic gradients are the fundamental physical parameter that determine movement of water in a porous media such as levees.

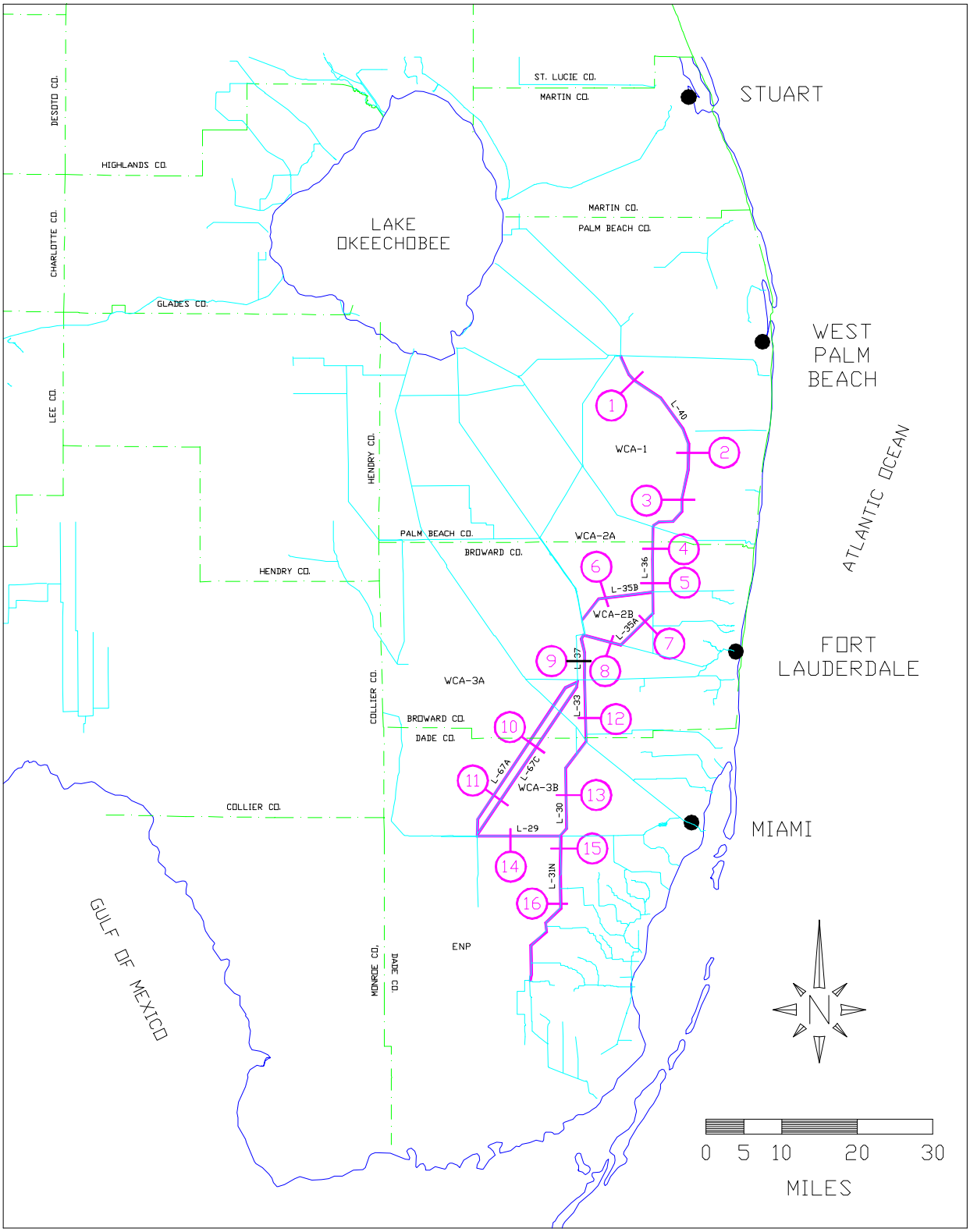

Figure 2.4.2 Sections or Transects Across the Major Levees Used to Formulate Levee Seepage Equations in the South Florida Water Management Model

4. Incorporate regression equations to SFWMM. This step involves the creation of a function which calculates [Eq. (2.4.2)] seepage volume from a grid cell to a canal located in an adjacent cell along the alignment of the N-S protective levee.

Using step-wise linear regression analysis, sixteen regression equations were established, one for each levee, relating levee seepage and prevailing head gradients. All equations were of the form:

$$Q_{seep} = \beta_0 + \beta_1 \Delta h_1 + \beta_2 \Delta h_2 \qquad (2.4.2)$$

where:

Q_{seep} = unit levee seepage, (cfs/mi);
$\beta 0, \beta 1, \beta 2$ = regression or levee seepage coefficients;
Δh_1 = head gradient across a levee representing the difference in the water levels inside a water conservation area and a levee borrow canal (local head gradient), (ft); and
Δh_2 = head gradient across a levee representing the difference in the water levels on opposite sides of a levee borrow canal (regional head gradient), (ft).

During the regression analysis, several cross-sections were found to produce very similar coefficients such that some of them were eventually grouped together and the analysis redone. Regression coefficients derived from this analysis were later referred to as levee seepage coefficients. Table 2.4.1 lists the ten sets of levee seepage coefficients used in the model. These coefficients were fine-tuned during model calibration.

Table 2.4.1 Levee Seepage Coefficients (β_0, β_1, β_2) Used in the SFWMM

Levee	β_0	β_1	β_2
L-40	0.1	1.3	-0.7
L-36	2.0	-1.0	1.8
L-35B	-0.2	1.0	-0.2
L-35A / L-35 / L-37	0.6	0.9	-0.2
L-67AC	-1.8	131.0	-128.3
L-33	0.5	15.5	-8.7
L-30	-8.4	170.6	-97.6
L-29	-0.4	53.5	-9.4
L-31N	1.0	75.0	-77.9
L-31	2.0	94.0	-77.9

Groundwater Flow

Governing Equations. Regional groundwater flow (or simply groundwater flow) in the SFWMM involves the solution of the partial differential equation (PDE) describing transient flow in a two-dimensional, anisotropic, heterogeneous, unconfined aquifer. The PDE is of the form:

$$\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_{yy}\frac{\partial h}{\partial y}\right) = S\frac{\partial h}{\partial t} - R \qquad (2.4.3)$$

where:

x and y = Cartesian coordinates aligned along the major axes of hydraulic conductivity or transmissivity;

T_{xx} and T_{yy} = transmissivity tensors of the aquifer, [L^2 T^{-1}];

h = the unknown hydraulic or potentiometric head, [L];

S = unconfined aquifer storage coefficient or specific yield of the porous media; vertically- averaged specific storage; volume of water release from storage per unit area of aquifer per unit decline in head, [dimensionless];

R = recharge; volumetric flux per unit surface area, [L T^{-1}]; and

t = time [T].

Equation (2.4.3) is strictly valid for confined aquifers only but is used in the model by allowing T_{xx} and T_{yy} to vary with time as saturated zone thickness changes (Wang and Anderson, 1982) since transmissivity is the product of hydraulic conductivity (assumed to be time-invariant in the model) and aquifer saturated thickness whose value varies as the location of the water table changes from one time step to the next. The derivation of the above equation makes the following assumptions:

1. Flow is essentially two-dimensional such that transmissivity, storage coefficient, recharge and hydraulic head can be vertically averaged.
2. The fluid, water, is incompressible.
3. Hydraulic conductivity, as well as transmissivity, is symmetric and the axes can be rotated such that the off-diagonal terms in the tensor are zero. In other words, the coordinate axes are assumed to be aligned with the major trends controlling hydraulic conductivity, in which, for example, flow in the x-direction is a result of the hydraulic gradient only in the x-direction. Thus, $T_{xy} = 0$ and $T_{yx} = 0$.
4. The momentum equation for an anisotropic medium is based on Darcy's Law which relates flow rate to an energy loss gradient by the hydraulic conductivity - Darcy's proportionality constant.
5. Drawdown or water table gradients are small relative to the saturated thickness.

Since the saturated thickness b is a function of hydraulic head h at any given time, the two-dimensional groundwater flow equation (2.4.3) is a nonlinear PDE. It is sometimes called the diffusion equation because it can be derived by performing a mass balance (continuity) and momentum balance (Darcy's Law) to describe the flow in a porous media.

In mathematical terms, Eq. (2.4.3) is classified as a parabolic partial differential equation that can be solved using a variety of numerical techniques. The SFWMM uses a variation of the Saul'yev method to solve the PDE given boundary and initial conditions (Saul'yev, 1964). The technique is unconditionally stable and explicit (direct) such that no iteration is required within a single time step. Initially, the region to be modeled is subdivided into a block-centered grid network with square and regular grid cells, i.e., $\Delta x = \Delta y$ = constant. The PDE is then transformed to its finite difference approximation: a system of linear algebraic equations written for each grid cell in the network. Lastly, the system of equations is solved sequentially until all nodal heads (average groundawater levels at grid cells) are determined.

Model Implementation. In order to minimize bias (or error propagation), the system of linear algebraic equations is solved in four different directions in four successive time steps. Unlike the overland flow subroutine, no time slicing is performed in the groundwater flow subroutine. A complete pass of all grid cells in the model domain is accomplished by one of the following directions:

1. left-to-right starting with the southwestern corner cell of the model domain, proceeding from the bottom row to the top row;
2. right-to-left starting with the northeastern corner cell of the model domain, proceeding from the top to the bottom row;
3. bottom-to-top starting with the southwestern corner cell of the model domain, proceeding from the left column to the right column; or
4. top-to-bottom starting with the northeastern corner cell of the model domain, proceeding from the right column to the left column.

Thus, four passes, one of each in the above sequence, through the grid network takes four time steps. Using Saul'yev method, the finite difference approximation of Eq. (2.4.3) is varied slightly depending on the direction by which the solution to the PDE is carried out. A basic derivation follows.

Consider the first term in Eq. (2.4.3). By taking the centered difference at the computational grid cell denoted by reference node (i,j) in terms of the midpoints and using a $0.5\Delta x$ spacing, we obtain

$$\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x}\right) \simeq \frac{1}{\Delta x}\left[\left(Txx\frac{\partial h}{\partial x}\right)_{i+1/2,j} - \left(Txx\frac{\partial h}{\partial x}\right)_{i-1/2,j}\right] \tag{2.4.4}$$

Expanding $\frac{\partial h}{\partial x}$ in both terms on the right side yields

$$\left[\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x}\right)\right]_{i,j} \simeq \frac{1}{\Delta x}\left[Txx_{i+1/2,j}\left(\frac{h_{i+1,j}-h_{i,j}}{\Delta x}\right) - Txx_{i-1/2,j}\left(\frac{h_{i,j}-h_{i-1,j}}{\Delta x}\right)\right] \tag{2.4.5}$$

where:

$Txx_{i+1/2,j}$ = transmissivity between node (i,j) and node (i+1,j); and

$Txx_{i-1/2,j}$ = transmissivity between node (i-1,j) and node (i,j).

The transmissivity terms can be evaluated at the midpoints of the grid cells. Three commonly used approximations are the arithmetic mean, geometric mean and harmonic mean, all of which produce satisfactory results for most groundwater flow problems (Willis and Yeh, 1987). Average conductivities or transmissivities in the horizontal direction are typically obtained using arithmetic means while those in the vertical direction are obtained using harmonic means. Non-directional averages are best estimated with geometric means. In the SFWMM, transmissivities are evaluated as arithmetic averages of transmissivities from adjacent nodes such that $Txx_{i+1/2,j}$ = 0.5 ($Txx_{i+1,j}$ + $Txx_{i,j}$) and $Txx_{i-1/2,j}$ = 0.5 ($Txx_{i,j}$ + $Txx_{i-1,j}$). If we let $T_{x1} = 1/\Delta x * Txx_{i-1/2,j}$ and $T_{x2} = 1/\Delta x * Txx_{i+1/2,j}$, we can simplify equality (2.4.5):

$$\left[\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x}\right)\right]_{i,j} \simeq T_{x1}\left[\frac{h_{i-1,j}-h_{i,j}}{(\Delta x)^2}\right] + T_{x2}\left[\frac{h_{i+1,j}-h_{i,j}}{(\Delta x)^2}\right] \qquad (2.4.6)$$

Using the same procedure for the y-derivative at node (i,j), we obtain

$$\left[\frac{\partial}{\partial y}\left(T_{yy}\frac{\partial h}{\partial y}\right)\right]_{i,j} \simeq T_{y1}\left[\frac{h_{i,j-1}-h_{i,j}}{(\Delta y)^2}\right] + T_{y2}\left[\frac{h_{i,j+1}-h_{i,j}}{(\Delta y)^2}\right] \qquad (2.4.7)$$

where:

$T_{y1} = 1/\Delta y * Tyy_{i,j-1/2}$ and $T_{y2} = 1/\Delta y * Tyy_{i,j+1/2}$.

Next, we take the forward difference approximation of $\frac{\partial h}{\partial t}$ relative to time t at the same reference node (i,j):

$$\left[\frac{\partial h}{\partial t}\right]^{t}_{i,j} \simeq \frac{h_{i,j}^{t+1} - h_{i,j}^{t}}{\Delta t} \qquad (2.4.8)$$

By evaluating all space derivatives in terms of time step t, a simple explicit formulation of PDE (2.4.3) results. However, some combinations of Δx and Δt in such a formulation result in numerical errors that could accumulate from one time step to the next, i.e., an explicit formulation is only conditionally stable. The Saul'yev method uses the computational efficiency of an explicit scheme while maintaining the stability of an implicit one. This method takes advantage of the direction of calculations in order to produce an explicit scheme based on an implicit formulation. For directions **1.** and **3.**, the solution proceeds in the +x and +y directions. Using Eqs. (2.4.6) through (2.4.8), the system of linear algebraic equations approximating PDE (2.4.3) takes the form:

$$T_{x1}\left[\frac{h_{i-1,j}^{t+1}-h_{i,j}^{t+1}}{(\Delta x)^2}\right] + T_{x2}\left[\frac{h_{i+1,j}^{t}-h_{i,j}^{t}}{(\Delta x)^2}\right] + T_{y1}\left[\frac{h_{i,j-1}^{t+1}-h_{i,j}^{t+1}}{(\Delta y)^2}\right] + T_{y2}\left[\frac{h_{i,j+1}^{t}-h_{i,j}^{t}}{(\Delta y)^2}\right]$$

$$= S_{i,j}\left[\frac{h_{i,j}^{t+1} - h_{i,j}^{t}}{\Delta t}\right] - R_{i,j}^{t+1} \qquad (2.4.9)$$

where:

$$\begin{aligned} T_{x1} &= \frac{T_{x_{i-1,j}} + T_{x_{i,j}}}{2} \\ T_{x2} &= \frac{T_{x_{i+1,j}} + T_{x_{i,j}}}{2} \\ T_{y1} &= \frac{T_{y_{i,j-1}} + T_{y_{i,j}}}{2} \\ T_{y2} &= \frac{T_{y_{i,j+1}} + T_{y_{i,j}}}{2} \end{aligned} \qquad (2.4.10)$$

Equation (2.4.9) is implicit because $h_{i-1,j}^{t+1}$, $h_{i,j}^{t+1}$, and $h_{i,j-1}^{t+1}$ appear simultaneously in the formulation. However, since the scheme proceeds in the +x and +y directions, then all values to the left (e.g., $h_{i-1,j}^{t+1}$) and below (e.g., $h_{i,j-1}^{t+1}$) the current cell (i,j) are known from a previous calculation during the same time step t+1. Thus, the unknown head $h_{i,j}^{t+1}$ is solved in terms of head values from the previous time step (old heads) and head values from the previous calculations (known or boundary heads) at the nodes surrounding (i,j).

Similarly, for directions **2.** and **4.**, the solution proceeds in the -x and -y directions and the corresponding system of linear algebraic equations to be solved is:

$$T_{x1}\left[\frac{h_{i-1,j}^{t} - h_{i,j}^{t}}{(\Delta x)^2}\right] + T_{x2}\left[\frac{h_{i+1,j}^{t+1} - h_{i,j}^{t+1}}{(\Delta x)^2}\right] + T_{y1}\left[\frac{h_{i,j-1}^{t} - h_{i,j}^{t}}{(\Delta y)^2}\right] + T_{y2}\left[\frac{h_{i,j+1}^{t+1} - h_{i,j}^{t+1}}{(\Delta y)^2}\right]$$

$$= S_{i,j}\left[\frac{h_{i,j}^{t+1} - h_{i,j}^{t}}{\Delta t}\right] - R_{i,j}^{t+1} \qquad (2.4.11)$$

Finally, we can solve for $h_{i,j}^{t+1}$ via:

$$h_{i,j}^{t+1} = \frac{(D + C - B * h_{i,j}^{t})}{A} \qquad (2.4.12)$$

where the following applies to directions **1.** and **3.**:

$$A = \frac{T_{x1}}{(\Delta x)^2} + \frac{T_{y1}}{(\Delta y)^2} + \frac{S_{i,j}}{\Delta t} \qquad (2.4.13)$$

$$B = \left[\frac{T_{x2}}{(\Delta x)^2} + \frac{T_{y2}}{(\Delta y)^2}\right] \tag{2.4.14}$$

$$C = \frac{S_{i,j}}{\Delta t} * h_{i,j}^{t-1} + R_{i,j}^{t} \tag{2.4.15}$$

$$D = \left[\frac{T_{x1} * h_{i-1,j}^{t+1} + T_{x2} * h_{i+1,j}^{t}}{(\Delta x)^2}\right] + \left[\frac{T_{y1} * h_{i,j-1}^{t+1} + T_{y2} * h_{i,j+1}^{t}}{(\Delta y)^2}\right] \tag{2.4.16}$$

while the following applies to directions **2.** and **4.**:

$$A = \frac{T_{x2}}{(\Delta x)^2} + \frac{T_{y2}}{(\Delta y)^2} + \frac{S_{i,j}}{\Delta t} \tag{2.4.17}$$

$$B = \left[\frac{T_{x1}}{(\Delta x)^2} + \frac{T_{y1}}{(\Delta y)^2}\right] \tag{2.4.18}$$

$$C = \frac{S_{i,j}}{\Delta t} * h_{i,j}^{t-1} + R_{i,j}^{t} \tag{2.4.19}$$

$$D = \left[\frac{T_{x1} * h_{i-1,j}^{t} + T_{x2} * h_{i+1,j}^{t+1}}{(\Delta x)^2}\right] + \left[\frac{T_{y1} * h_{i,j-1}^{t} + T_{y2} * h_{i,j+1}^{t+1}}{(\Delta y)^2}\right] \tag{2.4.20}$$

Figure 2.4.3 shows a typical computational grid used in the groundwater flow subroutine. The head at the grid cell denoted by (i,j) at time step t+1 is a function of the head at five adjacent cells, including itself, evaluated at time steps t and t+1. The selection of which cells to evaluate at a particular time step depends on the current direction of calculations.

Coupling of Groundwater and Surface Water

The solution to the governing groundwater flow equations [Eq. (2.4.3)] assumes a vertically homogenous, i.e., constant hydraulic properties, soil column and applies only to the saturated portion of the aquifer. The thickness of the saturated zone is assumed to be "unbounded" during the solution of the groundwater flow equations. If the water surface elevation goes above ground level, the assumption of homogeneity is violated: ponded water above land surface and saturated water in the aquifer have different hydraulic properties. The coupling of groundwater and surface water is further complicated by the existence of an intervening zone of aeration (unsaturated zone). The thickness of the unsaturated zone varies as the location of the water

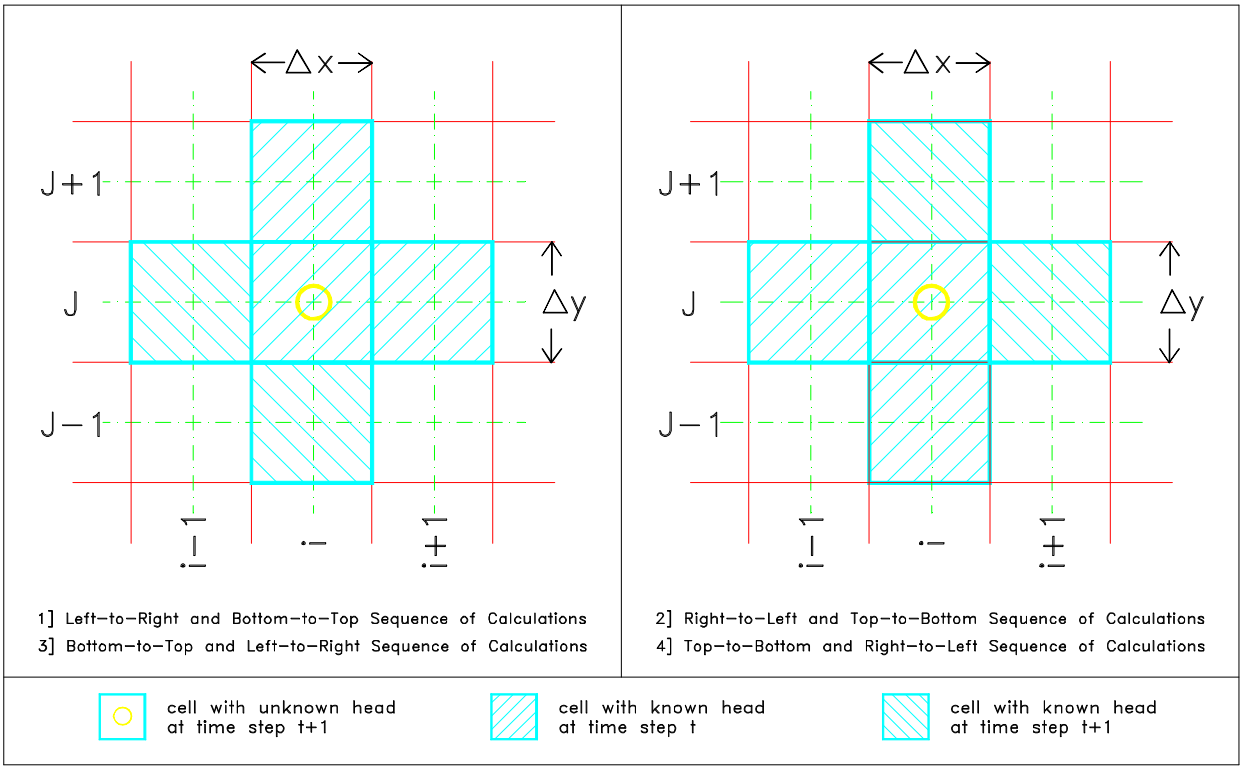

Figure 2.4.3 Location of Computational Cells Used in Calculating Total Head at Grid Cell (i,j) During Time Step t+1 as Implemented in the Groundwater Flow Subroutine in the South Florida Water Management Model

table fluctuates from one time step to the next. The model maintains mass balance for the unsaturated zone as its control volume changes with time. However, detailed physical processes such as lateral subsurface flow within this zone and capillary rise from the water table into the root zone are not modeled in the SFWMM.

Part of the algorithm used in the groundwater flow subroutine is the adjustment of the hydraulic or potentiometric heads just before and after the solution to the groundwater flow equations in order to account for differences in aquifer and ponded water hydraulic properties. These adjustments are often done in the wetland areas, e.g. WCAs and ENP, where occasional drying and rewetting of model grid cells occur. In these areas, the SFWMM assumes that dry soil conditions exist in the soil column above the water table and below land surface, i.e. the unsaturated zone is assumed not to exist. In the irrigated areas of the Lower East Coast region, an unsaturated zone moisture accounting procedure is performed (refer to Sec. 3.5). Moisture is assumed to be uniformly distributed within the unsaturated zone and always available for root uptake and plant transpiration.

A brief description of the variable names pertinent to the current discussion is as follows:

ells = elevation of land surface, [ft NGVD];

h = hydraulic or potentiometric head; elevation of groundwater; location of the

water table within the soil column relative to a datum, [ft NGVD]; For modeling purposes, this variable has a maximum value of land surface elevation.

infilt = infiltration; equivalent depth of water crossing land surface; typically water movement from ponding to the unsaturated zone, [ft];

perc = percolation; equivalent depth of water crossing the water table; typically water movement from the unsaturated zone to the saturated zone, [ft];

pond = depth of ponding, [ft];

S = storage coefficient for a confined aquifer; equivalent to the specific yield for an unconfined aquifer or the fraction (by volume) of water in a soil column released from (or gained into) storage per unit area of aquifer per unit decline (or increase) in head;

solmc = soil moisture content in the unsaturated zone, [ft];

t = time step, [day]; and

whc = water holding capacity in the unsaturated zone; equivalent to field capacity or the drained upper limit or the fraction (by volume) of water in a soil column above which water will percolate past the root zone and into the saturated zone.

Prior to the solution of the groundwater flow equations, if ponding exists, the known heads are reset to include the additional head provided by the ponded water, i.e.

$$h_t = h_t + pond_t \qquad (2.4.21)$$

A residual ponding term, whose value is equal to (1.0-S)*pond, is assumed not to take part in the solution to the groundwater flow equations. It is, however, be added back to the computed heads in order to maintain mass balance for each computational cell. The SFWMM also assumes that moisture in the unsaturated zone will not affect the solution to the groundwater flow equations. Moisture in this zone is updated at the end of the calculations if the computed heads encroach upon the unsaturated zone.

If the computed head, h_{t+1}, goes above land surface, the final ponding depth is updated to include residual ponding and/or unsaturated zone moisture content.

$$pond_{t+1} = (h_{t+1} - ells)*S + (1.0-S)pond_t + solmc_t \qquad (2.4.22)$$

The final head is equal to land surface elevation, i.e., $h_{t+1} = ells$.

If the computed head goes below land surface, residual ponding and unsaturated zone moisture are added back to the aquifer. Ponding depths and final heads are updated appropriately if the combined effects of residual ponding and unsaturated zone moisture content is to saturate the entire soil column. Otherwise, the final ponding depth becomes zero and the unsaturated zone mass balance is performed. For accounting purposes, additional infiltration and percolation will occur if residual ponding exist and percolation will increase if the water table encroaches upon the unsaturated zone to an extent that will bring the moisture content in this zone at field capacity.

One of the strengths of the SFWMM is its ability to simultaneously describe the state of the surface water and groundwater systems within the model domain. This state is defined in terms of ponding depths, unsaturated zone water content, and groundwater levels. The formulation of the recharge term (the combined effect of percolation, evapotranspiration, canal-groundwater seepage, and aquifer withdrawal for domestic, industrial and irrigation purposes) in the Eq. (2.4.3), levee seepage, and the procedure outlined in the preceding discussions comprise the vertical coupling of groundwater and surface water in the model.

Figure 2.4.4 shows a block diagram of the physical processes simulated in the model for surface and subsurface systems. Rainfall is a process that moves water from the atmosphere into surface storage. Evapotranspiration is the movement of water from both surface and subsurface systems into the atmosphere. A canal, which is essentially a special form of surface storage, exchanges water with ponding and the saturated zone storage through runoff/overbank flow and canal-groundwater seepage, respectively. Lastly, levee seepage is a localized flow phenomenon that describes the movement of water from the aquifer into a borrow canal across a major levee.

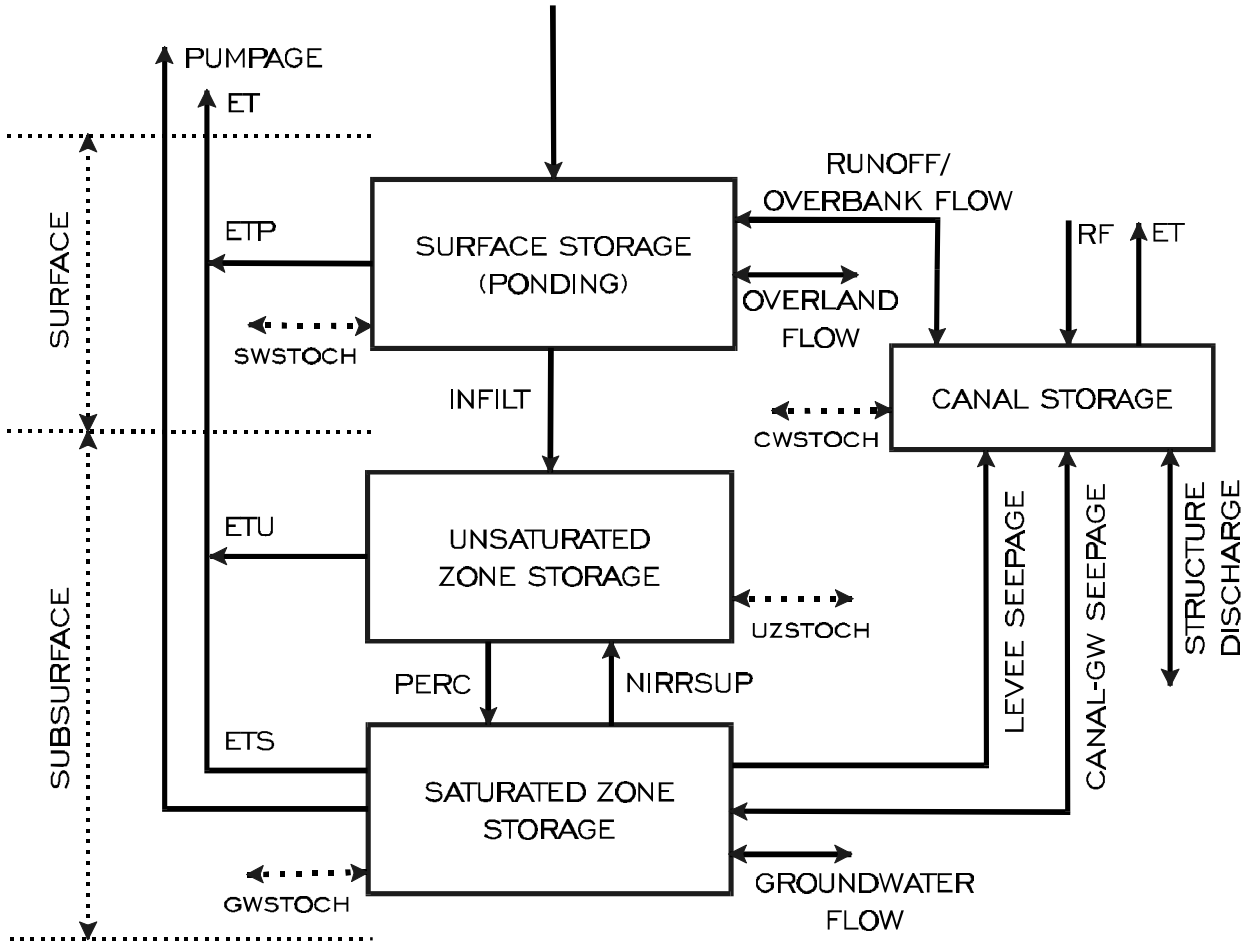

Figure 2.4.4 Generalized Block Diagram of Surface-Subsurface Interaction in the SFWMM